GIS-based recharge estimation by coupling surface–subsurface water balances

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Summary
A spatially distributed water balance model is developed to simulate long-term average recharge depending on land cover, soil texture, topography and hydrometeorological parameters. The model simulates recharge iteratively connected to a groundwater model, such that the recharge estimate is also influenced by the groundwater depth and vice versa. Parameter estimation for the model is performed on the basis of literature values of water balance fluxes from mainly Belgium and The Netherlands. By graphical and non-linear baseflow separation for 17 catchments it is shown that recharge spatially varies considerably. The water balance model coupled to a regional groundwater model is applied and successfully tested on the 17 catchments. The application shows that the resulting recharge has a spatial complex pattern, depending to a large extend on the soil texture and land cover. Moreover, shallow groundwater levels in valleys cause negative recharge conditions as a result of evapotranspiration by abundant phreatophytic vegetation. GIS analysis shows how recharge strongly varies for different combinations of land cover and soil texture classes. The performed analysis provides a better insight into the sustenance and management of groundwater resources.

Introduction
Recharge is the entry into the saturated zone of water made available at the water table surface, together with the associated flow away from the water table within the saturated zone (Freeze, 1969). Methods for determining recharge have been intensively discussed in literature (Simmers, 1988; Lerner et al., 1990; de Vries and Simmers, 2002; Scanlon et al., 2002). These can be classified as either point based or lumped areal methodologies. Cherkauer and Ansari (2005) present a lumped areal recharge estimation procedure in which regression relationships are established between recharge, as derived from baseflow, and catchment

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average topography, hydrogeology and land cover characteristics. Fazal et al. (2005) estimate recharge using a conceptual rainfall–runoff model, which requires only rainfall, evaporation and groundwater level data. Chloride mass-balance approach (Flint et al., 2002), isotopes (Guimerà and Candela, 1999; Scanlon et al., 2002), soil water balance modelling (Meiresonne et al., 2003; Ladekarl et al., 2005) and estimates on basis of groundwater level fluctuations (Sophocleous, 1991; Healy and Cook, 2002; Dickinson et al., 2004) are all point estimates of recharge. However, extrapolation or regionalization of point estimates is not evident (Sophocleous, 1992), and it is argued that distributed approaches as remote sensing based techniques offer chances for improvements in spatial estimates (Jackson, 2002). Regional mean annual distributed recharge is determined by Szilagyi et al. (2005) using the DRASIC methodology. Zhang et al. (1999) apply a biophysically based distributed ecohydrological model to simulate recharge for a small catchment. Armbruster and Leibundgut (2001) also use for a small area a distributed SVAT-model, however combined with a conceptual soil model to simulate temporal and spatial recharge. Cherkauer (2004) tests the deterministic, distributed parameter model PRMS for quantifying recharge per hydrological response units on a catchment scale of 50–500 km².

Regional groundwater models for analyzing groundwater systems (infiltration–discharge relations) are often steady state (Carrera and Medina, 1999) and therefore need long-term average recharge input. Also the spatial variation in the recharge, due to distributed land cover, soil texture, slope, etc., can be significant and should therefore be accounted for in these regional groundwater models. Chapman (1999) clearly shows the spatial variability of recharge by stream flow recession analysis.

Recharge rates are one of the most poorly constrained hydraulic parameters in almost all groundwater flow and transport models (Lerner et al., 1990; Anderson and Woessner, 1992). Therefore, in practice, recharge is often used as an adjustable parameter during model calibration (Anderson and Woessner, 1992; Sanford, 2002). However, because hydraulic conductivity can vary over 13 orders of magnitude (Freeze and Cherry, 1979), a practically infinite number of combinations of hydraulic conductivities and recharge rates that span a wide range could all match the target head distribution for a flow model calibration (Zhu, 2000). Consequently, inadequate control of recharge rates can lead to a situation of non-uniqueness of flow model solutions, which may render many water resource management models unusable (Brooks et al., 1994). However, Chardigny (1999) shows that, by coupling the aquifer and river mass balances, the overall calibration of a regional groundwater model could be improved.

The first motivation for this paper arose from the fact that in recent years many countries provided spatial variable land cover and soil data in digital form. Moreover, remote sensing furnishes us with a strongly increasing amount of high resolution spatial data. Therefore, it would be desirable to have an approach that directly takes into account the influence of the spatial variability of soil texture, land cover, slope and meteorological conditions in recharge estimation. Secondly, spatially distributed estimates of the recharge are needed for a better understanding of recharge—discharge systems and for a more reliable calibration of groundwater flow models. Thirdly, adopting sustainable groundwater use should lead to improved management of groundwater quantity and quality, but will be sub-ordinate to improved knowledge of spatial estimates of recharge (Sophocleous, 2005). Hence, the purpose of this paper is to develop a methodology which is flexible and open, i.e., GIS based, and consisting of a distributed water balance model coupled to a regional groundwater model for assessing spatially variable recharge. The results of the methodology will be tested and optimized with literature derived point estimates of water balance fluxes for similar hydro-meteorological and pedological conditions. Acceptance testing of the estimated water balance fluxes (Bogena et al., 2005) will be performed by comparing the water balances of 17 sub-catchments of the humid temporal Dijle, Demer, and Nete catchments in Belgium. Additionally, the distributed water balance is constrained by coupling to a regional groundwater model. Spatial statistics are used to describe and analyze the correlation between the water balance fluxes and the distributed land cover and soil maps, pointing to the need for a distributed modelling approach.

Methodology

A methodology for estimating spatially distributed, long-term average, recharge under humid temporal conditions was developed. This model, termed WetSpass, an acronym for Water and Energy Transfer between Soil, Plants and Atmosphere under quasi Steady State, integrates a water balance in a geographical information system (GIS) (Batelaan and De Smedt, 2001). The model has been applied to study the influence of long-term effects of land cover changes on the water regime in a basin (Batelaan et al., 2003; De Smedt and Batelaan, 2003). By using long-term average standard hydrometeorological parameters as inputs, the model simulates the temporal average and spatial differences of surface runoff, actual evapotranspiration, and groundwater recharge. Since evapotranspiration from shallow groundwater can be significant, especially in groundwater dependent wetlands, the position of the groundwater table should be taken into account in the estimation of recharge. Therefore, WetSpass is iteratively connected to a groundwater model, which provides the position of the water table, while WetSpass returns a recharge estimate accordingly.

The model treats a basin or region as a regular pattern of raster cells. Every raster cell is further sub-divided in a vegetated, bare soil, open water, and impervious surface fraction, for which independent water balances are maintained (Fig. 1). This allows to account for sub-cell land cover heterogeneity, depending on the resolution of the raster cell. Multi-resolution remote sensing classification offers possibilities to quantify sub-cell information (Van de Woerde et al., 2006). The bare soil fraction of a raster cell is also used to describe the part of the surface, which is not fully covered by vegetation. Especially, in the non-growing season this percentage can increase considerably for certain covers. The processes in each cell are simulated seasonally, while semi-temporal information is brought into the
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approach by assuming a cascading of the precipitation, interception, runoff, evapotranspiration and recharge processes.

The seasonal water balance for a vegetated fraction of a raster cell is

\[ P = I + S_v + T_v + R_v \]  (1)

where \( P \) is the precipitation [L], \( I \) the interception [L], \( S_v \) the surface runoff [L], \( T_v \) the actual transpiration [L] and \( R_v \) the groundwater recharge [L] in the vegetated fraction of the raster cell. The total actual evapotranspiration, \( ET_{tot} \) [L], is the sum of \( I \), \( T_v \) and the evaporation from the bare soil, \( E_s \) [L]. The actual evapotranspiration, \( ET_v \) [L], is the sum of \( T_v \) and \( E_s \). Interception has been shown to be reasonably constant at a given annual precipitation value and exhibits a consistent decrease with increasing annual rainfall total (Roberts, 1983). Therefore, the interception is parameterized as a constant fraction of precipitation depending on the vegetation type (Calder, 1979; Nonhebel, 1987; Dolman and Nonhebel, 1988; Grossmann, 1998).

The surface runoff, \( S_v \), is simulated in two stages. First, the potential surface runoff, \( S_{v-pot} \) [L], is calculated as

\[ S_{v-pot} = f_1(V, ST, S, D)(P - I) \]  (2)

where \( f_1 \) is a runoff factor from a lookup table whose value is dependent on vegetation type (\( V \)), soil texture (\( ST \)), slope (\( S \)) and groundwater saturated areas (\( D \)) and is based on characteristic values derived from Smedema and Ryckman (1988), Pilgrim and Cordery (1992), USDA-NRCS (1972), Chow et al. (1988). However, this potential surface runoff is conceptualized to occur only on groundwater saturated areas. Hence, in a second stage, the potential surface runoff is adjusted for recharge areas by taking into account the seasonal precipitation intensity distribution (\( P_i \)) in relation to the soil infiltration capacity (\( I_c \)) (Rubin, 1966).

\[ S_v = f_2(P_i, I_c, D)S_{v-pot} \]  (3)

where \( f_2 \) is a factor from a soil texture lookup table that partitions the precipitation for a hydrological season, in an effective and non-effective part for contribution to the surface runoff. It can be derived from estimating the fraction of seasonal precipitation with an intensity higher than the infiltration capacity of a particular soil type. It therefore characterizes the precipitation fraction contributing to the Hortonian surface runoff process for each soil texture. However, in groundwater discharge areas all potential surface runoff is assumed to contribute to surface runoff (Dunne and Black, 1970a,b), while in infiltration areas only high intensity storms will generate surface runoff.

Evapotranspiration is simulated with seasonal climatic data, essentially in a “top-down” approach, similar as in Zhang et al. (2001), however in this case not on a catchment but on a grid level. The reference or unstressed transpiration (Federer, 1979) is defined as

\[ T_{rv} = cE_o \]  (4)

where \( T_{rv} \) is the reference transpiration of a vegetated surface [L], \( E_o \) is the potential evaporation of open water [L], and \( c \) is a vegetation coefficient [–]. The vegetation coefficient can be determined as the ratio of the reference vegetation transpiration equation, as given by the Penman–Monteith equation, over the equation for Penman open water evaporation, and result in

\[ c = \frac{1 + (1 + r_v r_a)^{-1}}{1 + 1/(1 + r_v r_a)} \]  (5)

where \( I \) [ML\(^{-1}\)T\(^{-2}\)\,\,Θ\(^{-1}\)] is the proportionality constant or the slope of the saturation vapor pressure curve, \( \gamma \) [ML\(^{-1}\)T\(^{-2}\)Θ\(^{-1}\)] is the psychometric constant, \( r_v \) [TL\(^{-1}\)] is the aerodynamic resistance, and \( r_a \) [TL\(^{-1}\)] is the canopy resistance (Monteith, 1965). Dingman (2002) gives a similar vegetation coefficient as in Eq. (5), but includes a soil moisture dependent canopy resistance function from Stewart (1988).

In vegetated groundwater discharge areas, the actual transpiration, \( T_v \), is equal to the reference transpiration, since there is no limitation of soil water presence. In vegetated areas where the groundwater level is below the root zone the actual transpiration is determined as

\[ T_v = f(\theta)T_{rv} \]  (6)

where \( f(\theta) \) is a function of the water content (\( \theta \)) in root zone [–], \( G_d \) is the phreatic groundwater depth [L], \( h_i \) is the tension saturated height [L] (Brooks and Corey, 1964; Rawls et al., 1992) and \( R_s \) is the rooting depth [L] (Dickinson et al., 1993; Dingman, 2002; Famiglietti and Wood, 1994; Gehrels, 1999). For \( f(\theta) \) a formulation based on Vandevenie et al. (1991) and Xu (1992) is used

\[ f(\theta) = 1 - a\theta^{m} \]  (7)

where \( a \) is a calibrated parameter related to the soil texture, which increases with decreasing saturated hydraulic
conductivity, and \( w \) is the available water for transpiration [L]

\[
w = P + n(\theta_{tc} - \theta_{pwp})R_d
\]  

(8)

where \( P \) is the precipitation of a hydrological season consisting of \( n \) [-] months, \( \theta_{pwp} \) and \( \theta_{tc} \) are respectively the water content in root zone at permanent wilting point [-] and at field capacity [-], and \( \theta_{tc} - \theta_{pwp} \) is the plant available water content.

The soil classification for the WetSpass model is based on the classification of the US Department of Agriculture, Soil Survey Staff (1951). For each soil texture clay and sand fractions have been determined as midpoint values of each textural class (Cosby et al., 1984). With these fractions the field capacity, permanent wilting point and plant available water content are calculated based on generalized soil water characteristic equations (Saxton et al., 1986). Comparison of these parameter values with Rawls et al. (1992), Dingman (2002), ASCE (1990), and Kabat and Beekma (1994) shows a good correspondence.

Change in storage is introduced in the model by allowing different groundwater levels for winter and summer conditions and by assuming that during winter the plant available soil moisture reservoir is filled up so that during summer this reservoir can be depleted.

Groundwater recharge can be determined as the residual of the water balance (Eq. (1)). The presented methodology results in an estimation of spatially distributed recharge as a function of vegetation, soil texture, slope, groundwater depth and precipitation regime. Even in groundwater discharge areas some recharge will be simulated, which is in agreement with the conceptual idea that a thin unsaturated zone is generally still present in discharge areas, allowing some recharge. However, in summer the simulated recharge in discharge areas will often be negative as a result of the high transpiration of the vegetation due to the presence of shallow groundwater.

A procedure for the bare soil surfaces can be developed in a similar manner as for the vegetated surfaces. Recharge is determined from the water balance for bare soil surfaces as

\[
P = S_s + E_s + R_s
\]  

(9)

where the index \( s \) refers to bare soil surfaces. The surface runoff, \( S_s \), is simulated in a similar way to the vegetated area fraction in two stages (Eqs. (2) and (3)). The bare soil evaporation is determined as

\[
E_s = f(\theta)E_o
\]  

(10)

where \( f(\theta) \) is a factor, which is defined by Eq. (7) and in which \( E_v \) is replaced by \( E_{ps} \), a Penman evaporation rate for a wet soil.

The water balance for the open water area fraction of a cell is defined as

\[
P = E_o + S_o + R_o
\]  

(11)

where the index \( o \) stands for open water surfaces. It is assumed that the recharge, \( R_o \), derived from the precipitation on the open water fraction, is negligible compared to the possible recharge from the surface water body itself. If the Penman open water evaporation, \( E_o \), is smaller than the precipitation, the remaining will contribute to the surface runoff, \( S_o \).

The water balance for impervious surfaces is given as

\[
P = S_i + E_i + R_i
\]  

(12)

where the index \( i \) refers to impervious surfaces. Due to the wide range of possible types of impervious surfaces and their physical characteristics, the percentage of surface runoff and evaporation has to be specified by the user based on knowledge and observations of the characteristics of these surfaces.

The total water balance, per raster cell and hydrological season, can now be stated as

\[
E_{tc} = a_vE_v + a_sE_s + a_oE_o + a_iE_i
\]  

(13)

\[
S_c = a_vS_v + a_sS_s + a_oS_o + a_iS_i
\]  

(14)

\[
R_c = a_vR_v + a_sR_s + a_oR_o + a_iR_i
\]  

(15)

where the index \( c \) refers to raster cell, with \( E_{tc}, S_c, R_c \) [L] respectively, the total evapotranspiration, surface runoff and recharge in a raster cell and \( a_v, a_s, a_o \) and \( a_i \) respectively the vegetated, bare soil, open water and impervious area fractions of a raster cell.

The simulated distributed recharge can now serve as input for a groundwater model, and the resulting groundwater level or depth can consequently return as input for WetSpass as illustrated in Fig. 2. In general, iteration between the groundwater model and WetSpass will converge in a few iterations (Batelaan et al., 2003). The WetSpass and regional groundwater model are spatially coupled and parameters are estimated jointly. Hence, it is more important to assure that the water balance is spatially realistic as opposed to blindly optimizing model parameters over the

![Figure 2](image_url)  

**Figure 2** Schematic representation of the iterative process between WetSpass and a groundwater model.
entire basin in order to match one stream flow gauge near the outlet of the basin (Arnold et al., 2000).

WetSpass is tightly integrated in ArcView and ArcGIS and coded respectively in Avenue and Visual Basic. The GIS raster model structure is very appropriate since it enables to connect GIS easily with a numerical groundwater model as well as with input data derived from satellite imagery. GIS data structures are used efficiently since spatial information is stored in raster layers while parameter information is stored in attribute tables. From the attributes new raster layers can be derived to use in spatial calculations. The attribute tables also allow to define easily new land cover or soil types, as well as changes in the parameter values, which permits analysis of future land and water management scenarios.

Parameter estimation with point fluxes

Parameter estimation of the WetSpas methodology is performed on the basis of literature values of interception, evapotranspiration and recharge studies from mainly Belgium and The Netherlands. These literature values are compared to simulated WetSpas water balance fluxes for typical combinations of land cover, soil texture, average topographic and long-term meteorological conditions of Ukkel, Belgium. WetSpas input data for this test are summarized in Table 1. The groundwater depth is set at 4 m, in order to simulate infiltrating conditions. The simulation results for these climatological conditions and all possible combinations of typical land cover types and the 12 USDA soil textures are given in Fig. 3. The figure shows for each land cover the simulated average and standard deviation of the surface runoff and the recharge, while for interception, transpiration and (soil) evaporation only the average is given as a stacked bar diagram.

The effectively determined interception by WetSpas depends on the parameterized winter and summer interception percentages, but also on the Penman wet-surface evapotranspiration and the fraction of soil surface covered by vegetation. The resulting effective interception percentages for different vegetation types, presented in Fig. 4, compare well with values obtained from literature (Molchanov, 1960; Leyton et al., 1967; Schnick, 1971; van Roestel, 1984; Nonhebel, 1987; de Visser and de Vries, 1989; Hendriks et al., 1990; Moors et al., 1996; Meiresonne et al., 1999). In Nonhebel (1987), simulated forest stands gave consistently higher interception values with the Mulder model (Mulder, 1985) than with the Gash model (Gash, 1979). Moors et al. (1996) summarize the interception measurements for forests in the Netherlands as 20% of precipitation (26% in summer, 10% in winter) for deciduous forests and 39% of precipitation for non-deciduous forests. WetSpas calculates very similar interception values. Further confirmation of the WetSpas interception values is provided by van Roestel (1984), who finds for needle forest a range of 30–40%, for deciduous forest 20–30% in hydrological summer and 10–20% in hydrological winter. Shuttleworth (1992) indicates that for deciduous forests, the loss of leaves in winter reduces the fractional interception loss typically by a factor of 2–3, which is in accordance with the parameterized interception values. The interception losses for short crops are lower than for forests, primarily because of increased atmospheric transport of water vapor from their aerodynamically rough surfaces (Calder, 1998).

Summarizing, it can be concluded that interception simulated with WetSpas corresponds well with values cited in literature. The resulting interception values range from about 15 to 325 mm for the following land covers, cited in increasing order of interception: crops and urban areas, shrub, heather, grass, orchard, mixed and deciduous forest (poplar, oak, birch, beech), coniferous (spruce, pine).

The simulated average transpiration (Fig. 3) for the different land cover types has a low standard deviation, indicating that the transpiration is not very dependent on the soil texture, as was also shown by Zhang et al. (2004). de Visser and de Vries (1989) also noted small differences in transpiration of about 50 mm due to different soil textures, Nonhebel (1987) reported a difference of 20 mm. Simulated transpiration for oak is 306 mm/yr, which is close to the range 277–293 mm as reported by Nonhebel (1987), Meiresonne et al. (1999) calculated the transpiration for a popular stand in East Flanders during four years as 278 mm/yr, using the soil water balance model WAVE (Van Clooster et al., 1994). For the same soil texture the transpiration calculated with WetSpas was 267 mm/yr. Nonhebel (1987)

<table>
<thead>
<tr>
<th>#</th>
<th>Soil</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Precipitation&lt;sup&gt;a&lt;/sup&gt;</td>
<td>400 mm (summer), 380 mm (winter)</td>
</tr>
<tr>
<td>2</td>
<td>Penman ET&lt;sup&gt;b&lt;/sup&gt;</td>
<td>543 mm (summer), 114 mm (winter)</td>
</tr>
<tr>
<td>3</td>
<td>Windspeed&lt;sup&gt;c&lt;/sup&gt;</td>
<td>3.84 m/s (summer), 3.27 m/s (winter)</td>
</tr>
<tr>
<td>4</td>
<td>Temperature&lt;sup&gt;d&lt;/sup&gt;</td>
<td>14.1 °C (summer), 5.0 °C (winter)</td>
</tr>
<tr>
<td>5</td>
<td>Land cover&lt;sup&gt;e&lt;/sup&gt;</td>
<td>20 Types</td>
</tr>
<tr>
<td>6</td>
<td>Soil&lt;sup&gt;f&lt;/sup&gt;</td>
<td>12 Standard USDA textures</td>
</tr>
<tr>
<td>7</td>
<td>Slope</td>
<td>&lt;0.5%</td>
</tr>
<tr>
<td>8</td>
<td>Groundwater depth</td>
<td>4 m (infiltration condition)</td>
</tr>
</tbody>
</table>

<sup>a</sup> Average Ukkel 1833–1975 (Dupriez and Sneyers, 1978; Sneyers and Vandiepenbeek, 1995).
<sup>c</sup> Long-term average Ukkel (Sneyers and Vandiepenbeek, 1995).
<sup>d</sup> OC GIS-Vlaanderen (1996).
<sup>e</sup> Original Belgian soil map (Van Ranst and Sys, 2000; OC GIS-Vlaanderen, 2001) converted to textures of USDA (1951).
simulated the transpiration of a pine stand in the range of 176–189 mm, while the transpiration simulated with WetSpass is 189 mm. Moreover, WetSpass simulates an average transpiration rate of 327 mm/yr for different needle and deciduous forests, corresponding to Roberts (1983), who found an average transpiration value of 333 mm/yr for several needle and deciduous forests in Western Europe, while van Roestel (1984) gives a range of 270–370 mm for the yearly transpiration of forests in moderate climates.

Soil evaporation was simulated by Meiresonne and Van Slycken (1999) as 88 mm on average for a poplar stand. The corresponding WetSpass simulated value is 91 mm. Soil evaporation simulated by de Visser and de Vries (1989) for forests, heather, grassland and bare soil in The Netherlands, also compares well with the corresponding WetSpass values.

Gehrels (1999) simulated recharge with a soil water balance model for a period of more than 30 years for different forest stands in The Netherlands. These recharge values show an increase from dark over light coniferous to deciduous birch forest, as respectively 129, 331 and 362 mm/yr. WetSpass test results indicate a similar trend with comparable values of respectively 123, 316 and 349 mm/yr. In addition, the model shows that recharge steadily increases from pine over deciduous forests to grassland. This is also confirmed by Schroeder (1983) with long-term lysimeter data of Castricum in The Netherlands and St. Arnold in Germany, showing that the sum of surface runoff and recharge from grassland is respectively 100 and 220 mm/yr larger than for deciduous and pine forest. Querner (2000) simulated recharge for pasture, arable, pine and deciduous land covers on sandy soils as respectively 336, 338, 217 and 333 mm/yr. The corresponding values calculated with WetSpass are respectively 344, 337, 285, 317 mm/yr, which closely correspond to the values given by Querner (2000).

This good agreement between point estimates of the water balance fluxes and WetSpass simulated values is not the result of an a priori ‘upward’ physical approach. On the contrary, it is rather the result of a ‘downward’ process in which model concept and parameters, starting from a very simple water balance approach, have been adapted step by step, while continuously verifying model results with literature and independent runoff data (Sivapalan et al.,

Figure 3  Simulated average (○) of (a) surface runoff, (b) interception, transpiration, evaporation, and (c) recharge for typical land cover and forest stands calculated for Ukkel climatological conditions. The vertical bar in (a) and (c) indicates plus or minus one standard deviation due to the different soil types. For (b) only the averages are given.
The soils in the Dijle, Demer and Nete catchment are presented in Fig. 5 (Van Ranst and Sys, 2000; OC GIS-Vlaanderen, 2001). The main texture types are sand (30%), loam-sand (24%) and silt (21%). The area is characterized by a gradual change in texture from North-West to South-East, with a sand (24%) and silt (21%) component. The area is characterized by a gradual change in texture from North-West to South-East, resulting in a sequence of sand, loamy-sand and silt regions. In the Dijle basin the major soil textures are sandy-loam and loamy-sand. In the Demer basin the complete sequence of sand, loamy-sand, sandy-loam to silt occurs from North to South. The transition region with sandy-loam and loamy-sand textures becomes narrower towards the East. The Nete catchment is dominated in the South and in the valleys by loamy-sand and in the North-West by soil types with a sand texture.

In Fig. 6 a simplified version of the land cover map for the Dijle, Demer and Nete catchments is given. The actual land cover map has 22 classes and is a result of a supervised maximum likelihood classification of Landsat 5, Thematic Mapper (TM) images of 5, 12 and 19th of August 1995 (OC GIS-Vlaanderen, 1996). The classification result was improved by using the CORINE land cover (NGI, 1994), hydrographical, road and soil association maps. The study area is dominated by agriculture (35%), forests (24%), built-up areas (20%) and meadows (14%).

Precipitation has been measured in Brussels (Ukkel) since 1833. No significant trend in precipitation was found (Vaes et al., 2002). Dupriez and Sneyers (1978) determined the statistics of this time series, as well as for 360 stations of the Belgian national network for the period 1951–1975. On the basis of these statistics, monthly and yearly isohyetal precipitation maps were interpolated with respect to the period 1833–1975 (Dupriez and Sneyers, 1978). The spatial interpolation has an error of maximum 5% for stations most distant from Ukkel. The precipitation increases from 693 mm/yr in the southwest of the Demer catchment up to 866 mm/yr in the north east of the Nete catchment. The spatial average precipitation for the total area is 773 mm/yr (st.dev. 23 mm/yr). Seasonal precipitation differences are limited to a 3% higher precipitation in summer (April till September) than in winter (October till March). The average yearly potential open water evaporation ranges from 662 to 675 mm/yr. The summer potential evaporation typically constitutes about 85% of the total yearly amount. The average windspeed is 3.5 m/s and almost constant for the winter and summer season, while the temperature is respectively 5.0 and 14.1 °C for the winter and summer (Sneyers and Vandiepenbeeck, 1995).
Figure 5  Soil textures of the Dijle, Demer and Nete catchments (Van Ranst and Sys, 2000; OC GIS-Vlaanderen, 2001).

Figure 6  Land cover of the Dijle, Demer and Nete catchments.
In total 17 river gauging stations, 5 in Dijle, 8 in Demer and 4 in Nete catchment, were selected for hydrograph analysis (Fig. 7), each with 10 years or more of daily discharge data available. The watersheds belonging to the gauging stations were derived from a digital elevation model, based on contour levels of a 1:100,000 scale map, and vary in size between 6 and 2260 km², as shown in Fig. 7.

The WetSpas model was set-up for the Dijle, Demer and Nete catchment with a grid resolution of 50 × 50 m. Additionally, a MODFLOW groundwater flow model was developed with the same resolution (Batelaan and De Smedt, 2004). This model was extended with a new MODFLOW package for a better simulation of seepage areas, as discussed by Batelaan and De Smedt (2004).

Results

Baseflow analysis

Despite the conceptual problems and negative opinion towards the use of hydrograph separation tools (Hewlett and Hibbert, 1967; Appleby, 1970; Nathan and McMahon, 1990; Tallaksen, 1995; Chapman, 1999, 2003), the technique remains popular and justified (Dingman, 2002) to estimate spatial averaged recharge. Traditionally, graphical techniques to separate the hydrograph in baseflow and surface runoff are used. More recently, these techniques have been automated by applying algorithms to systematically identify baseflow of a stream hydrograph (Nathan and McMahon, 1990; Sloto and Crouse, 1996; Rutledge, 1998), e.g., Sloto and Crouse (1996) presented the HYSEP program in which fixed interval, sliding interval, and local minimum algorithms are used. Wittenberg (1999) developed a non-linear storage-discharge relationship equivalent to the differential non-linear power recession formulation based on solutions of the Dupuit–Boussinesq hydraulic theory (Brutsaert and Nieber, 1977; Troch et al., 1993).

The baseflow for the 17 stations is separated with the methods of Sloto and Crouse (1996) and Wittenberg (1999). The comparison of the estimated baseflow percentages reveals a correlation coefficient of $r = 0.64$. A non-parametric sign test shows that there is a significant difference ($p < 0.01$) between the average baseflow percentage of all stations obtained with the Sloto and Crouse (1996) and Wittenberg (1999) methodology. Wittenberg (1999) noted that the flexibility of the non-linear baseflow separation technique yields a computed baseflow closer to the total discharge hydrograph than can be achieved with classical linear approaches. A non-parametric Kruskal–Wallis test shows significant differences ($p < 0.01$) in the baseflow between the catchments. These differences can be observed in Fig. 7, which shows the spatial distribution of the baseflow percentages for the 17 sub-catchments calculated with the Sloto and Crouse (1996) methodology. The parts of the Dijle, Demer and Nete River catchments belonging to the more downstream stations show a lower baseflow percentage. This can be explained by the shallow groundwater conditions and higher urbanization downstream, causing higher runoff and evapotranspiration and consequently reducing the baseflow in the more downstream parts of the catchments.

Figure 7 Baseflow percentages in different sub-catchments of Dijle, Demer and Nete, determined by the discharge separation technique of Sloto and Crouse (1996).
Distributed water balance

The simulated runoff ranges from 1 to 636 mm/yr. The average runoff is 49 mm/yr as observed from the areal cumulative distribution in Fig. 8, while for Dijle, Demer and Nete this is respectively 56, 53 and 42 mm/yr. Fig. 9a gives the synopsis of the origin of the runoff. It is striking that about 55% of the total runoff is produced on impervious surfaces, while much smaller contributions are obtained from vegetated, bare soil and open water surfaces.

Fig. 10 shows the estimated yearly evapotranspiration. The map shows a complex distributed pattern, with values ranging from 212 to 675 mm/yr. The average is 465 mm/yr for the total study area (Fig. 8), and respectively 475, 473 and 453 mm/yr for Dijle, Demer and Nete. The components of the evapotranspiration in the different catchments are summarized in Fig. 9b. On yearly basis the total evapotranspiration consists for about 55% of transpiration, while bare soil evaporation, interception and evaporation from impervious and open water surfaces play a much smaller role, that is respectively 19%, 17%, 6% and 2%. On a seasonal basis the picture looks quite different. In summer the transpiration comprises 68%, while in winter this is only 19% of the total evapotranspiration. On the other hand, bare soil evaporation contributes only 8% in summer, while this increases to 53% in winter.

Fig. 11 shows the yearly groundwater recharge estimated with WetSpass. The values range from –384 to 462 mm/yr; the average amounts to 251 mm/yr (Fig. 8) and for Dijle, Demer and Nete this is respectively 236, 231 and 280 mm/yr. Fig. 9c shows that the recharge from bare soil is a major component in the total recharge. The recharge in the Dijle and Demer catchment occurs for more than 99% during the winter. In the Nete, however, only 87% of the yearly recharge occurs in the winter, due to the influence of the sandier soils in this catchment.

Figure 8  Areal cumulative distributions of simulated runoff, evapotranspiration and recharge fluxes for combined Dijle, Demer and Nete area.

Distributed water balance

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Figure 9  Areal averages for Dijle, Demer and Nete catchments of WetSpass simulated: (a) surface runoff; (b) evapotranspiration; and (c) recharge fluxes.
Figure 10  
WetSpass simulated yearly evapotranspiration in Dijle, Demer and Nete catchments.

Figure 11  
WetSpass simulated yearly recharge in Dijle, Demer and Nete catchments.
Linked groundwater and surface water balance modelling

The set-up and calibration of the groundwater model has been extensively described by Batelaan and De Smedt (2004). Here only the results are given of the coupling of the model with WetSpass for simulation of the water balance in Dijle, Demer and Nete catchments, and comparison with measured river discharges in the 17 sub-catchments. Fig. 12a and b shows scatter plots of respectively measured versus simulated total discharge and measured baseflow versus simulated recharge. The correlation coefficients ($R^2$) for a regression with intercept 0 are respectively 0.95 and 0.93, which indicates that the agreement between model results and observations is very good. In order to evaluate the model performance, five global error measures were used to compare the simulated and measured total discharge: the relative mean and mean absolute error are respectively 10% and 19%, the Nash–Sutcliffe model efficiency (Nash and Sutcliffe, 1970) is 0.89, and the coefficient of determination is 0.69. All these error measures indicate that the fit between observations and model estimates is very good.

Discussion

Notwithstanding the good comparison between the model results and the observations, obviously differences remain. These errors can be attributed to uncertainties in the complexity of the model-data space, i.e., model concept, observation errors and uncertainties in land cover, soil classifications, slope and precipitation maps. For some catchments the assumed correspondence of topographic and groundwater divides might not be correct. Also, the topographic catchment divide might not be representative for the actual river catchment, because artificial transfers of water by channels, ditches, sewer systems and pumped groundwater can disturb the natural water balance. Another shortcoming is the non-representativeness of present day soil and land cover conditions to simulate long-term average water balances.

In order to understand the spatial recharge distribution, it is useful to first analyze evapotranspiration and surface runoff, which occur prior to the recharge. From Fig. 10 it is observed that in the southern part of the Dijle and Demer catchments the average evapotranspiration is high, while in the northern part of the Dijle and Demer catchments and in the Nete catchment this is lower. The southern high values are the result of silty soil types, which are able to hold more soil water and yield more evapotranspiration. This is also the case for the high evapotranspiration values in the (wet) valleys, which usually consist of more clayey type of soils. Especially, in the Demer basin the difference between valleys and surrounding areas is obvious, due to silty valley soils in contrast to loamy sand and sandy loam soils on the interfluves. On the eastern rim of the Nete and Demer catchment several coniferous forests are clearly noticeable by their relatively high evapotranspiration.

Differences in evapotranspiration between the basins, shown in Fig. 9b, indicate that the total evapotranspiration for Dijle and Demer is very similar, while it is about 5% lower in the Nete basin. This difference occurs mainly in summer, since in winter evapotranspiration is very small, as solar energy is limited. In summer, energy is not the limiting factor, but rather the water availability. This explains why the Nete, with sandier soils, has a lower transpiration and bare soil evaporation than the other two catchments. On the other hand, the interception and open water evaporation in the Nete catchment is about 30% higher than in Dijle and Demer. This is mainly due to the higher forest cover, which is 29.0% in Nete, 24.6% in Dijle, and 19.9% in Demer catchment.

It would be interesting to reveal the correlation between the evapotranspiration map and one or more of its related factors, as soil, land cover, precipitation, etc. The covariance of one map with another can be measured and described quantitatively, however the significance of a particular correlation value in a probabilistic sense is much more difficult to evaluate with spatial data than with non-spatial data. Instead of using the term ‘significant’, it is safer to talk about ‘unusual’ or ‘interesting’ correlations, drawing the attention to an association, rather than

Figure 12  Comparison of observed total yearly average discharge (a) and baseflow (b) with WetSpass simulation results.
implying that a correlation can be rejected at some level of probability (Bonham-Carter, 1994).

Also, global correlations between mapped variables are usually not very high, and correlations often appear locally only under a given set of conditions. Bonham-Carter (1994) and Whitman et al. (1999) describe a method to quantize the mapped variables into a finite number of discrete classes and then compare the degree of overlap between classes by means of a cross-area tabulation. This results in a Chi-square statistic, which has a lower limit of 0 and a variable upper limit. Its value is further dependent on the unit of measurement. Therefore, Cramer’s coefficient V is often used instead (Eastman, 1999). V varies from 0, no correlation between maps, to 1, perfect correlation.

The maps of precipitation, potential evaporation, total actual evapotranspiration, runoff and recharge are sliced in intervals of 5 mm/yr, while the slope is sliced in intervals of 0.25%, and the groundwater depth in intervals of 0.5 m. The Cramer’s V correlation coefficients determined with the GIS IDRISI32 (Eastman, 1999) are given in Table 2. This reveals that evapotranspiration has a high, interesting, correlation with land cover. The soil and potential open water evaporation maps show lower correlations, while the other maps are less correlated with evapotranspiration. Runoff shows an interesting association with land cover, but also with soil texture, while other maps are less correlated. On the other hand, recharge is only moderately correlated with land cover and soil texture. Overall, land cover seems to have the highest impact on the distribution of the different water balance components, while soil texture is the second most important. The slope has very little impact, which can be explained by the rather gentle topography of the study area. The moderate impact of the precipitation and potential evapotranspiration can be explained similarly, i.e., these show relatively little variation over the study area. In the lower part of Table 2, correlations are also given between the WetSpass output maps. It follows that these correlations are similar and moderate, however evapotranspiration has slightly higher correlations with the output maps, obviously an effect of the influence of the land cover, as discussed above.

Since the land cover and soil map display the highest correlation with the resulting WetSpass output maps, it is useful to investigate the means and standard deviations of the evapotranspiration, runoff and recharge for the different land cover and soil texture classes of the Dijle, Demer and Nete catchments. Fig. 13 presents the mean and standard deviation of WetSpass simulated runoff, evapotranspiration and recharge per land cover and soil texture class for the total study area. It is noticeable from this figure that land cover has a stronger influence on evapotranspiration than soil texture. The evapotranspiration also has smaller standard deviations for each land cover class, which implies less impact of other factors. Most of the variability in the evapotranspiration for a particular land cover is due to differences in soil texture. Therefore, Fig. 14, the mean evapotranspiration is given for each combination of land cover and soil class, together with recharge and runoff. In this figure, less frequent combinations, as land cover with clay loam and clay soil texture, are left out. Notice that the evapotranspiration values increase about 75 mm/yr within each land cover class going from sandy to silty soil, likewise runoff increases markedly. These increases of evapotranspiration and runoff with finer soil textures are compensated by a reduction of recharge values. Fig. 14 shows also that there are large differences in runoff between on the one hand natural and agricultural landuses and on the other hand the urbanized areas, which have runoff values of factor 2–5 larger.

Overall, the large range in runoff values can be explained by the large impact of the land cover and soil texture (Fig. 13) on the generation of runoff. Intermediate runoff, i.e., 50–200 mm/yr, is mainly produced by urban areas, roads, etc., while large runoff, more than 200 mm/yr, occurs on open waters and groundwater saturated areas. In the Demer and especially the Nete catchment, the lowest runoff values occur on sandy soil textures in the northeastern part of these catchments. Although the correlation between on the one hand runoff and on the other land cover and soil textures is by far the most important (Table 2), local influences of slope and precipitation can be noticed. The seasonal difference in the runoff is relatively small, although the runoff is slightly higher in summer than in winter. This is mainly due to higher runoff in summer from vegetated surfaces; this effect is reduced in winter due to higher runoff from bare soil and open water.

Fig. 9c indicates that the recharge on bare soils is an important component of the total recharge. However, bare soil should be understood as the parts of grid cells with a land cover class like deciduous forest or agricultural fields, which especially in winter have a large non-vegetated (bare) cover. Hence, since the leaf area is low in the non-growing season and moreover the precipitation intensity, transpiration and evaporation are low, the recharge in these bare soil parts of otherwise vegetated areas is high. Since recharge is a result of evapotranspiration and runoff processes it incorporates all influences and spatial patterns of these processes. This causes a complex spatial distribution pattern of the recharge (Fig. 11). Negative recharge values are found in the valleys, especially in Dijle and Demer, but constitute less than 2% of the total area. These areas are groundwater discharge locations with high runoff and high evapotranspiration by abundant phreatophytic vegetation, resulting in negative recharge, which is balanced by the inflowing groundwater. On the other hand, high recharge rates, i.e., larger than 400 mm/yr, also occur in less than 1.5% of the area. These are located close to the northern and

<table>
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<th>Recharge</th>
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north-eastern water divides of the Nete and Demer basin on sandy soils and mostly maize land cover. Furthermore, recharge is lower than average in the southern part of the Dijle and Demer catchments, especially in several valley and urban locations. Similarly, in the western part of the Demer catchment, lower recharge values occur due to higher evapotranspiration. Larger than average recharge values are found in the sandy soil region in the northeastern part of the Demer and Nete catchment. This results in larger recharge values for the Nete catchment than for the two other catchments. Fig. 9 shows that higher recharge for the Nete is not the result of a relatively low evapotranspiration, but of a lower runoff compared to the two other catchments. This again can be attributed to soil texture, but also in particular to less built-up areas in the Nete (22%) compared to the Demer (24%) and especially Dijle (34%) catchment.

Overall, the spatial distribution of recharge is not specifically related to either land cover, soil or any other input maps, as can be observed from the relatively small Cramer’s V coefficients (Table 2). The highest correlation for the recharge map is given by the soil map. Fig. 13 shows that the different soil textures cause a somewhat larger variation in recharge values than the land cover classes, which explains the higher correlation between the recharge and soil texture map than with the land cover map. Fig. 14 shows that recharge decreases markedly with finer soil texture irrespective of the land cover classes. Agriculture, grassland and heather result in high recharge only on sandy soils. On the other hand, deciduous and mixed forests show also relatively high recharge but their dependence on soil texture is clearly less significant. Although Fig. 14 allows a good interpretation of the relative importance of recharge per land
cover and soil combination, the total areal recharge contribution is more important for groundwater management. Therefore, in Fig. 15 recharge per land cover and soil class combination is shown as a pie chart, together with the percentage areal coverage of each class and the contribution of each land cover class to the total recharge in the study area. From this figure it follows that maize has the highest areal coverage, as well as an even higher contribution to the total recharge, i.e., the ratio of the recharge over the area percentages is larger than one. From a groundwater management point of view, this land cover type is most important for groundwater resources sustenance. However, this is also often accompanied by a high nutrient load, with a high risk for pollution of the groundwater resources. Other land cover types with a high impact on the total recharge compared to their areal percentage are the different forest covers and heather. Agriculture, meadow and wetlands have recharge contributions that are lower but close to their areal coverage. All built-up classes have very low relative recharge contributions. In view of their considerable areal coverage (20%), these land covers do not contribute significantly to the total recharge, and hence, have a high potential for improving the recharge conditions.

In the used methodology, the linking of groundwater and water balance (recharge) modelling is essential. From the recharge map (Fig. 11) it is evident that simulated shallow groundwater tables have a marked influence on the evapotranspiration, runoff and resulting recharge in the valleys. It is estimated that transpiration from groundwater occurs in 26% of the total study area. This evapotranspiration from groundwater contributes to only 2% of the total evapotranspiration, including interception, of the study area. However, the groundwater transpiration rises to an average of 11% of the total transpiration in areas where groundwater influences the transpiration process. Very locally this increases up to 90%. In 6% of the groundwater affected transpiration zones or 1.6% of the total area, i.e., mainly in some of the broader valleys, groundwater contributes 50% or more to transpiration. As a consequence of the higher transpiration and also higher surface runoff on these partly groundwater saturated zones, recharge is about 40 mm lower than outside the zone of groundwater transpiration. Hence, it is essential to take the groundwater levels into account in the estimation of the spatial distribution of the recharge in order to avoid more uncertainty in optimizing hydraulic conductivities by inverse modelling. Moreover,
the use of the MODFLOW evapotranspiration package is not needed since the present methodology also accounts for the evapotranspiration losses from groundwater.

The advantage of embedding the methodology in GIS is that it allows easy evaluation of the effects of land cover changes on recharge. Additionally, the GIS approach proves to be very useful in the spatial analysis of the simulated recharge, and could be used to optimize the recharge conditions and improve groundwater protection and sustainability. Moreover, the GIS structure of the model with parameters as attribute tables has been shown to be advantageous in transferring it to other environments. The model is modest in its data requirements and has the possibility that remote sensing derived data layers can be used. WetSpass was successfully applied in Belgium as well as in other type of environments like in the Upper-Biebrza catchment, Poland (Thijs, 2002) and the Geba Basin, Tigray, Ethiopia (Asfaw, 2005).

Conclusions

Spatial distribution of recharge is seldom taken into account in groundwater simulations, although poorly parameterized recharge can increase significantly the uncertainty in modelling results and calibration, especially the estimation of the hydraulic conductivity. Baseflow separation techniques for the Dijle, Demer and Nete catchments indicate a significant spatial distribution of the recharge, which proves the need for distributed spatial simulation of recharge. With the advent of Geographic Information Systems and the availability of spatial and remotely derived data sets, a logical step is to estimate recharge as a result of a surface water balance simulation. The developed GIS-based WetSpass methodology is therefore a tool that, in conjunction with a groundwater model, can simulate accurately the spatial distribution of long-term average recharge. The development of the model and the parameter estimation is not the result of an a priori ‘upward’ physical approach, but rather a ‘downward’ process, starting from a simple water balance approach, which subsequently evolved by introducing relevant concepts and input data step by step (Sivapalan et al., 2003). The model results correspond well with point estimates of water balance fluxes in Belgium and The Netherlands. The combined WetSpass and groundwater model was applied to the Dijle, Demer and Nete catchments and on the basis of five global error measures was acceptably verified with measured discharges and estimated baseflows in 17 sub-catchments. The model concept can relatively easy be transferred to other areas due to its open GIS based concept.

In about 25% of the study area shallow water table conditions cause transpiration from groundwater, on average...
11% of total transpiration. Recharge in these areas is reduced by 15% compared to the overall average. However, with shallow groundwater conditions and abundant vegetation, the recharge can locally reduce to negative values, indicating the importance of taking the simulated groundwater level into account in a recharge estimation procedure. Although these seepage areas with negative recharge values constitute less than 2% of the total area, they give structure to the simulated complex spatial distribution of recharge in the Dijle, Demer and Nethe catchments. Cramer’s Chi square statistics show that although the recharge pattern has no particularly high correlation with any input map, soil texture and land cover have the most impact. However, simulated evapotranspiration and surface runoff show higher correlations with land cover. Surface runoff is for about 55% produced on impervious surfaces, which is 2–5 times higher than runoff from natural and agricultural land covers. Total evapotranspiration consists for about 55% of transpiration. Recharge appears to be strongly seasonal, i.e., almost exclusively occurring in winter, except in the sandy Nethe catchment where 13% of the recharge occurs in summer. The Nethe catchment also has a recharge, which on average is considerably higher than in the other catchments. This is caused by a lower runoff, which is due to a flatter terrain, sandy soils, and less built-up area. Recharge on bare soils as a part of vegetated areas, i.e., deciduous forest and agricultural fields in winter, is shown to be an important component of total recharge. GIS analysis of recharge per land cover shows that agriculture, grassland and heather on sand produce the highest recharge, while deciduous and mixed forests also produce high recharge, but are less dependent on the soil texture. Obviously, impervious surfaces exhibit the lowest recharge values. Taking the areal surface of a land cover into account, maize fields contribute the most to the total recharge. Hence, from groundwater quantity management point of view, maize fields are very important for the groundwater resources. However, maize fields on the other hand are a high risk with respect to groundwater pollution by nutrients and threaten therefore severely the sustainability of the groundwater resources. Likewise, positively contributing to the groundwater resources are forest and heather. Other agriculture, meadow and wetlands are average recharge contributors. Urbanized areas in combination with their 20% areal coverage are the worst with respect to recharge, but as such have the highest potential for improving recharge conditions. The presented GIS-embedded methodology to estimate recharge has the significant advantage that it allows optimization of recharge conditions and improvement of groundwater management.

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